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Atmospheric Modeling and Its Spatial Representation of Land Surface Characteristics

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Atmospheric modeling is a technique used to study atmospheric phenomena in a mathematical and physical framework. The behavior of the atmosphere can be described using a set of differential equations that describe external forcing to the system and the response of the atmosphere to that forcing. In order to solve the equation set, initial and boundary conditions must be provided. However, there is currently no comprehensive geographical data set of land surface characteristics (e.g., canopy structural and radiometric properties, soil hydrological properties) that can provide surface boundary conditions for these models. GIS have become a natural choice for the reconciliation and storage of such data available from different sources with different projections and spatial resolutions.

In the following sections, we briefly describe atmospheric models and provide an overview of three current land surface parameterization schemes and their data requirements. Finally, using observations and atmospheric model simulations, we demonstrate the need for accurate characterization of the land surface.

ATMOSPHERIC MODELS

Atmospheric models range in spatial scales from the entire globe to one kilometer or less. The time period of model simulations range from on the order of an hour for the smallest domain size simulations to centuries for global simulations. Each of these models, regardless of domain size, is based on conservation laws of physics. As summarized in Pielke (1984), these relations are expressed as:

- Conservation equation for velocity;
- Conservation equation for heat;
- Conservation equation for air;
- Conservation equations of water substance in its three phases;
- Conservation equations of other chemical constituents.

The conservation relationship for velocity is obtained from Newton's second law of motion, while the conservation equation for heat is from the first law of thermodynamics. The conservation equations permit an accounting of the changes in each of the variables in time, and are expressed as a simultaneous set of partial differential equations.

Mathematically these equations can be expressed as

$$\frac{d\varphi}{dt} = \frac{\partial\varphi}{\partial t} + \vec{V}\cdot\vec{\nabla}\varphi = S_\varphi \qquad (10\text{-}1)$$

where φ can represent dependent variables such as velocity, potential temperature, water substance, etc. In this equation, $\vec{V}$ is the three-dimensional velocity vector. The source–sink term S_φ represents the nonconservative physical mechanisms by which these dependent variables can be changed. For velocity, for example, the pressure gradient force and gravitational force represent terms in S_φ. For potential temperature, radiative flux divergence, heat changes due to phase changes of water substance, etc. represent S_φ.

In the creation of a numerical model, this simultaneous set of partial differential equations is integrated over grid volumes corresponding to the spatial separation within the model. The result is a set of partial *difference* equations that have correlation terms to represent subgrid-scale processes. Mathematically, this integration can be represented as

$$\frac{\partial \overline{\varphi}}{\partial t} = -\vec{\nabla} \cdot \vec{\nabla}\overline{\varphi} - \left(\overline{\overline{\vec{V}} \cdot \vec{\nabla}\varphi''} + \overline{\vec{V}'' \cdot \vec{\nabla}\overline{\varphi}} + \overline{\vec{V}'' \cdot \vec{\nabla}\varphi''} \right) \tag{10-2}$$

where the overbar represents the grid volume average and the terms in parentheses represent the subgrid-scale fluxes. The dependent variable with the double primes represents the subgrid-scale values of the variable.

It is important to recognize that in meteorological numerical models, because the equations are discretized, only a portion of these models are explicitly based on fundamental physical quantities. These are:

- The pressure gradient force;
- Advection;
- Gravitational force.

The Coriolis force is also a fundamental concept, but it only appears because the meteorological equations are defined with respect to the rotating earth.

The quantities that are *parameterized* (and, in general, are not necessarily based on fundamental concepts) are:

- Cumuliform clouds;
- Stratiform clouds;
- Long- and short-wave radiative flux divergence;
- Subgrid-scale turbulent and coherent circulation fluxes;
- Soil and vegetation effects.

There are specific named subsets of atmospheric numerical models. These include, from smaller to larger domain size:

- Large eddy simulation (LES) models;
- Boundary layer (BL) models;
- Cumulus cloud models;
- Cumulus field models;
- Mesoscale models;
- Regional models;
- Hemispheric models;
- Global models.

The smallest-scale models (LES, BL, and cumulus cloud and cumulus field models) usually use the more general version of the conservation equations. Mesoscale and large-scale models often utilize the hydrostatic equation to substitute for the vertical velocity conservation equation. Regional, hemispheric, and global models are often used routinely for numerical weather prediction (NWP). For instance, the U.S. National Weather Service currently creates prognosis meteorological fields twice daily using a regional model referred to as the Nested Grid Model (NGM). The European Centre for Medium Range Weather Forecasting (ECMWF) performs model forecasts out to 10 days or so using the ECMWF global forecast model. General circulation models (GCMs) are a form of global models that up to the present have been integrated using coarser model resolution than the NWP global models. A recent summary of GCM modeling techniques is presented in Randall (1992).

The term *resolution* has been used in a wide variety of peer-reviewed publications to refer to the grid increment used in a model. For example, general circulation models (GCMs) are said to have a resolution of about 400 km by 400 km when that scale more appropriately refers to the horizontal grid mesh.

From sampling theory it is well known, however, that at least two grid increments are required to represent data. Real information at scales smaller than two grid increments is erroneously aliased to larger scales. An illustration of this is presented in Pielke (1984, Figure 10.7). Models such as GCMs, however, require additional grid resolution to simulate meteorological processes adequately as a result of serious computational inaccuracies at scales less than four grid increments (e.g., see Tables 10.1, 10.2, and 10.3 in Pielke, 1984). Some investigators suggest even more grid increments are needed for adequate simulations.

Using these clarifications, resolution within a numerical model should refer to at least four times the grid interval. For instance, a GCM with 400 km by 400 km horizontal grid increments would have a resolution of no less than 1600 km by 1600 km. Diagnostic data (e.g., terrain) with sampling at a 400 km interval would have a resolution of no better than 800 km. Information that has spatial resolution smaller than the resolvable scales, however, is not completely lost. Using statistical techniques, some of this information can still be captured. This is commonly referred to as "subgrid-scale parameterization," as briefly discussed earlier in this section. Subgrid-scale land surface heterogeneities, for instance, should be parameterized so that the integral effect on the atmosphere of these subgrid-scale land surface features can be accounted for.

Atmospheric motion is fueled by energy received from the sun, and a majority of this energy is first absorbed by the surface and then transferred to the atmosphere through surface turbulent exchange processes. Consequently, atmospheric processes are sensitive to surface characteristics. For example, a simple sensitivity analysis (Pielke et al., 1993) showed that a small change of surface albedo can result in a change of equilibrium atmospheric temperature as large as the proposed greenhouse warming effect. Given human ability to alter surface characteristics, we expect that landscape changes will have a large impact on global climate and weather. In the next section we overview three current land surface parameterizations that are used in atmospheric models to describe surface exchange processes. In the long term, we expect to upgrade these surface schemes to simulate the response of land cover (i.e., vegetation) to changes in atmospheric conditions.

OVERVIEW OF LAND PARAMETERIZATION SCHEMES USED IN NUMERICAL MODELS

Since the Earth's surface is the only natural boundary of the atmosphere, to include a surface representation scheme in numerical weather/climate prediction models is not a new idea. In his famous work on numerical weather prediction, Richardson (1922) noted:

> The atmosphere and the upper layers of the soil or sea form together a united system. This is evident since the first meters of ground have a thermal capacity comparable with 1/10 that of the entire atmospheric column standing upon it, and since buried thermometers show that its changes of temperature are considerable. Similar considerations apply to the sea, and to the capacity of the soil for water.

Richardson then went on to discuss the possible treatments of three principal surface covers, namely the sea, bare soil surface, and vegetation covered surface. For land surface, Richardson considered the motion of water in soil, the transfer of heat in soil, and evapotranspiration. Analogous to the electric conductance, the rate of transpiration is proportional to the stomatal conductance and the vapor pressure difference between the intercellular space and the canopy air. He described the physics in the soil–vegetation–atmosphere continuum as:

> Leaves, when present, exert a paramount influence on the interchanges of moisture and heat. They absorb the sunshine and screen the soil beneath. Being very freely exposed to the air they very rapidly communicate the absorbed energy to the air, either by raising its temperature or by evaporating water into it. A portion of rain, and the greater part of dew, is caught on foliage and evaporated there without ever reaching the soil. Leaves and stems exert a retarding friction on the air.

It has been 70 years since Richardson published his book on numerical weather prediction, and the idea of treating the exchange processes in the soil–vegetation–atmosphere continuum is still the same. A similar modeling concept is still widely used except that many of the detailed physical and biophysical processes have become understood over the years. New model parameters have been introduced. For example, the relation between the leaf area index (LAI) and the stomatal conductance (Jarvis and McNaughton, 1986) and between the root zone water stress and the evapotranspiration (ET) (Kramer, 1949) are parameterized in recent land surface models. It should be noted that even the ideas of introducing the LAI and the root zone water stress were briefly mentioned in Richardson's book.

Three land surface parameterization schemes, which are currently used in numerical weather/climate prediction models, are overviewed in this paper. The simple biosphere model (SiB; Sellers et al., 1986) and the biosphere–atmosphere transfer scheme (BATS; Dickinson et al., 1986) are used in general circulation models (GCMs). The land ecosystem–atmosphere feedback model (LEAF; Lee, 1992) is used in a smaller-scale model (the regional atmospheric modeling system; RAMS).

Model structure

As mentioned in the previous section, each of the land surface parameterization schemes currently used in atmospheric models have virtually the same model structure. Following Richardson's "vegetation-film" and Deardorff's (1978) "big-leaf" concept, the three models (BATS, LEAF, and SiB) each introduce a layer of vegetation that interacts with the atmosphere. Notice that although SiB has a separate layer of ground cover vegetation, it is treated as part of the ground (no prognostic equations are introduced to describe the thermal and hydrological properties of this layer). Besides, the recent simplified simple biosphere model (SSiB; Xue et al., 1991) has eliminated the second layer of vegetation and still obtains nearly the same results as SiB.

For this layer of vegetation, averaged quantities, such as wind speed, thermal capacity, exchange coefficients, and radiative extinction coefficient, are utilized so that the detailed flow structure and the interception of radiation in the canopy are not resolved. However, prescribed wind profile and radiation distributions are used to calculate these averaged quantities. This type of model has been classified as a "greenhouse canopy" model by Goudriaan (1989) so that, in addition to the "big leaf," the impact of the canopy air is also parameterized. Figure 10-1 shows schematically the structure of the three models. In this figure and the following sections, the notation used in SiB (Sellers et al., 1986) is also used for BATS and LEAF for ease of discussion. The variables H and λE are sensible and latent heat fluxes, T is temperature, e is water vapor pressure, Ψ is water potential, and r indicates a resistance function. Subscripts r, a, g, c, and s denote variables at different locations, namely the atmospheric reference level, the canopy air at the zero plane displacement height, the ground-level vegetation coverage, the canopy, and the soil surface, respectively. Three aerodynamic resistance functions are used: $\bar{r}_b$ is a bulk boundary layer resistance,

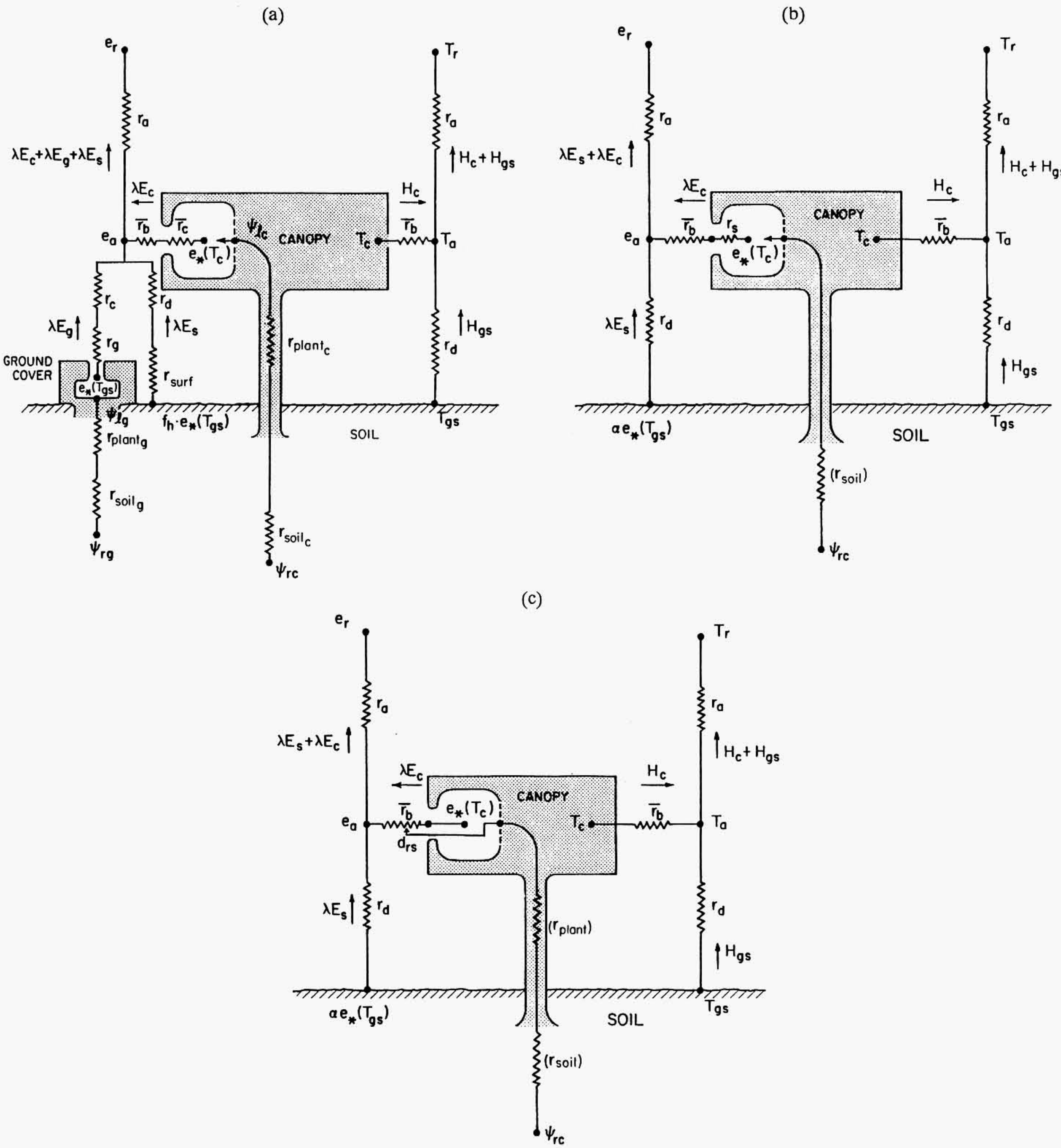


Figure 10-1:(a) Framework of the simple biosphere(SiB) model. The transfer pathways for latent and sensible heat flux are shown on the left- and right-hand sides of the diagram, respectively (from Sellers et al., 1986). Symbols are described in the text. (b) same as (a) but for the biosphere–atmosphere transfer scheme (BATS). (c) same as (a) but for the land ecosystem atmosphere feedback model (LEAF).

which is a resistance function between leaves and the canopy air, r_a is a resistance function between canopy air and the atmospheric reference level, and r_d is a resistance between the ground and the canopy air. In Figures 10-1b and c, the resistance functions in parentheses indicate the functions are implicitly embedded in other functions.

For example, in Figure 10-1c the resistance functions for water transport between the roots and the canopy, that is, r_{plant} and r_{soil}, implicitly exist in the formulation of relative stomatal conductance d_{rs}. The bulk stomatal resistance is noted as $\bar{r}_c$ in these figures.

It is not too difficult to see that these models have a similar structure, as shown in Figure 10-1. The canopy and the ground first exchange sensible and latent heat with the surrounding air and then the canopy air exchanges heat with the boundary layer atmosphere. We view this two-step exchange process as a major improvement to the "big-leaf" model in that the surface heat and moisture fluxes from the vegetation to the atmosphere are regulated by the heat and moisture capacity of the canopy air. Also, the water flow from the root zone to the surface of the leaves is regulated by the soil resistance, the plant resistance, and the bulk stomatal resistance. Although the three models show a similar model structure, it is the parameterization of these resistance functions that are different. A detailed comparison between the models is presented in the following sections.

Aerodynamic resistance

The primary goal of the existing soil–vegetation models is to provide a realistic boundary forcing to the atmosphere. This is accomplished through parameterizing the momentum, sensible heat, and latent heat fluxes. A common practice in parameterizing these fluxes is to utilize the resistance formulation:

$$\text{flux} = \frac{\text{potential difference}}{\text{resistance}} \qquad (10\text{-}3)$$

Employing the Monin–Obukhov surface-layer similarity theory, the momentum flux (or the shear stress) in all three models is in the form of:

$$\tau = \rho C_D u^2 = \rho u_*^2 = \rho \left[\frac{ku}{\ln\left(\frac{z-d}{z_0}\right) + \Psi_M} \right]^2 \qquad (10\text{-}4)$$

where τ is the shear stress, ρ is the density of air, u is the wind speed at a reference height z within the surface layer, C_D is the drag coefficient, u_* is the friction velocity, k is von Karman's constant, d is the zero plane displacement height, z_0 is the surface aerodynamic roughness, and Ψ_M is a stability adjustment function for momentum transport. In this formulation, we have assumed that the average flow speed at the displacement height is zero. The displacement height, d, and the surface roughness length, z_0, are prescribed in LEAF and BATS and are calculated in SiB by using "K theory." An additional assumption must be made, in order to calculate the displacement height, about the variation of the momentum transfer coefficient in the canopy. Sellers et al.(1986) noted that using "K theory" in the canopy may not be realistic and it also introduces new parameters. However, additional flexibility is obtained by using the SiB formulation so that the displacement height and surface roughness are actually varying with the density of the canopy (i.e., the LAI). This is important especially in climate simulations since the LAI varies with season. It may not be physically consistent to hold d and z_0 fixed while varying the LAI, such as applied, for example, in BATS. For short-range weather forecasts, where LEAF has been used, this may not be a problem, and we can supply these variables as model parameters and hold them fixed during the period of the model simulation (usually no more than several days).

For the sensible and latent heat fluxes, the boundary layer aerodynamic resistance function, r_a, is used in LEAF and SiB, so that:

$$H = \rho c_p \frac{T_r - T_a}{r_a} \qquad (10\text{-}5)$$

$$\lambda E = \rho \frac{c_p}{\gamma} \frac{e_r - e_a}{r_a} \qquad (10\text{-}6)$$

$$\frac{1}{r_a} = \frac{k^2 u}{\left[\ln\left(\frac{z-d}{z_0}\right) + \Psi_M\right]\left[\ln\left(\frac{z-d}{z_0}\right) + \Psi_H\right]}. \qquad (10\text{-}7)$$

In these formulas, the variable Ψ_H is a stability adjustment function for heat and moisture transport, λ is the latent heat of vaporization, and γ is the psychrometric constant. The stability adjustment functions can be found, for example, in Paulson (1970), Businger et al., (1971), and Louis (1979). Paulson's scheme is used in SiB, while Louis' scheme is used in LEAF and SSiB due to the fact that the latter scheme is noniterative. A slightly different form for the boundary layer resistance function is used in BATS where

$$\frac{1}{r_a} = u \left[\frac{k}{\ln\left(\frac{z-d}{z_0}\right)} \right]^2 \Phi, \qquad (10\text{-}8)$$

in which Φ is a stability correction function suggested by Deardorff (in Dickinson et al., 1986).

The bulk boundary layer resistance, $\bar{r}_b$, should depend upon the morphology of the vegetation and take into account the bluff-body effect of air flow around the leaves. Following the work by Goudriaan (1977), LEAF and SiB use the form:

$$\frac{1}{\bar{r}_b} = \int_0^{\mathrm{LAI}} \frac{\sqrt{u_f}}{C_s P_s} d\,(\mathrm{LAI}) \tag{10-9}$$

where u_f is the wind speed in the canopy, C_s is a transfer coefficient that depends on the shape of the leaves, and P_s is a shelter coefficient. As noted by Sellers et al.(1986), the major difficulty in using this formula is the determination of the shelter coefficient, P_s. This coefficient is highly dependent on the morphology of the vegetation and can only be determined empirically (e.g., Thom, 1972). A much simpler form is used in BATS, where

$$\frac{1}{\bar{r}_b} = \mathrm{LAI}\, C_f \left(\frac{u_f}{D_f}\right)^{\frac{1}{2}}, \tag{10-10}$$

in which C_f is an exchange coefficient and D_f is the dimension of the leaves in the wind flow. Although this formula is quite similar to the one used by LEAF and SiB, the effect of P_s on the bulk aerodynamic coefficient is not taken into account. This might be a drawback of this formula; however, on the other hand, the shelter coefficients, P_s, is one of the most ill-defined coefficient in LEAF and SiB due to the lack of measurements.

Knowing the bulk boundary layer resistance, the resultant sensible heat flux from the leaves to the canopy air can be written as:

$$H_c = \rho c_p \frac{T_a - T_c}{\bar{r}_b}, \tag{10-11}$$

where T_a and T_c are the temperature of the canopy air and the leaves, respectively. The resultant latent heat flux will be discussed later with the stomatal resistance function.

The surface aerodynamic resistance, r_d, should also depend on the morphology of the vegetation. For example, a constant stress profile may be able to be used in a hardwood forest, where the wind profile may be logarithmic near the surface and below the elevated canopy. However, there may be no turbulent exchange of heat and moisture between the soil and the canopy air beneath a dense grass canopy since the wind reduces to zero in the canopy. For sparse canopy, it is even more difficult to parameterize this effect. The relative contribution from the soil should be very important when the density of the canopy is small.

Assuming the logarithmic wind profile is valid beneath an elevated canopy layer, Sellers et al., (1986) used the following form in SiB:

$$r_d = \frac{C}{u_f \varphi_H} \tag{10-12}$$

where C is a surface-dependent constant and u_f is an average wind speed between the soil surface and the displacement height, and φ_H is a stability correction factor. The variable C can be calculated from a prescribed wind profile law. As mentioned, this formulation is designed for hardwood forests, but would be expected to fail when applied to other vegetation cover where the canopy is not elevated.

Due to the fact that this soil-to-canopy exchange process is not yet clearly understood, BATS chooses to use a simple formula to represent this effect without a fundamental physical base. Following Deardorff's (1978) work, a linear interpolation of resistance values between bare and vegetated surfaces is used, so that:

$$\frac{1}{r_d} = \frac{1}{r_a}\left[(1 - \sigma_f) + \frac{u}{u_f}\sigma_f \right] \tag{10-13}$$

where σ_f is the fractional coverage of vegetation. Recently, Dickinson (1991, personal communication) indicated that a change has been adopted to make the transition between vegetation and bare soil smoother. It is observed, in this formula, that r_d approaches r_a when vegetation coverage is small and $r_a u_f / u$ when the coverage is large. Following Shuttleworth and Wallace (1985), a similar form is used in LEAF, with some variation:

$$r_d = r_{\mathrm{bare}} \max\left(\left(1 - \frac{\mathrm{LAI}}{3}\right), 0\right) + r_{\mathrm{close}} \min\left(\frac{\mathrm{LAI}}{3}, 1\right) \tag{10-14}$$

where r_{bare} is the resistance function when the surface is bare and r_{close} is the resistance when the surface is covered by closed canopy. Instead of vegetation cover, LEAF uses LAI as an indicator of the coverage. It is assumed that there is no soil contribution when LAI is larger than 3. The advantage of using this formula is that it uses a realistic wind profile law to calculate the resistance functions, r_{bare} and r_{close}. The resultant turbulent heat fluxes from the soil are:

$$H_{gs} = \rho c_p \frac{T_a - T_{gs}}{r_d}, \tag{10-15}$$

$$\lambda E_{gs} = \rho \frac{c_p}{\gamma} \frac{e_a - \alpha e_*(T_{gs})}{r_d}, \tag{10-16}$$

where $e_*(T_{gs})$ is the saturation water vapor pressure immediately above the surface with a surface temperature T_{gs}. The variable α is an adjustment factor (or surface resistance) in determining the soil surface water vapor pressure.

Stomatal resistance

The major difference between vegetated and bare soil surface is the access to water in the soil. Over a bare soil surface, water is available for evaporation only from the top soil layers. In the presence of vegetation, water is also available from deep soil layers where roots are present. Although the transfer of water in the plant is mostly passive (meaning that water is not directly used by the photosynthesis process), it is responsible for the transport of nutrition from the root zone to the leaves where the photosynthesis process is taking place. The result of this transport process is that water is lost to the atmosphere through the opening (stomata) on leaves. It is known that water vapor pressure is at its saturation value in the intercellular space (Rutter, 1975). However, the vapor pressure at the surface of the leaves is regulated by the size of opening of the stomata, which is, in turn, a function of the environmental variables (e.g., photosynthetically active radiation, water stress, temperature, and CO_2 concentration).

Assuming this stomatal opening can be parameterized by a single resistance function, r_s, the water vapor flow from the intercellular space to the canopy is a two-step process. First, water vapor is transferred to the leaf surface:

$$\lambda E_1 = \rho \frac{c_p}{\gamma} \frac{[e_{\mathrm{sfc}} - e_*(T_c)]}{\bar{r}_c} \tag{10-17}$$

where $\bar{r}_c$ is a canopy resistance function, or bulk stomatal resistance function, e_{sfc} is the water vapor pressure at the surface of the leaves, and $e_*(T_c)$ is the saturation water vapor pressure at the intercellular space with the temperature of the leaves, T_c. Following Jarvis and McNaughton (1986), $\bar{r}_c$ is defined as:

$$\frac{1}{\bar{r}_c} = \int_0^{\mathrm{LAI}} \frac{1}{r_s} d(\mathrm{LAI}) \tag{10-18}$$

Second, water vapor is transferred to the canopy air:

$$\lambda E_2 = \rho \frac{c_p}{\gamma} \frac{(e_a - e_{\mathrm{sfc}})}{\bar{r}_b}. \tag{10-19}$$

If we further assume there is no accumulation of water vapor at the surface of the leaves (i.e., $E_1 = E_2$), we obtain:

$$\lambda E_c = \rho \frac{c_p}{\gamma} \frac{(e_a - e_*)}{\bar{r}_b + \bar{r}_c}. \tag{10-20}$$

In this equation E_c is the evaporation rate from the intercellular space to the canopy air. This equation is used in both BATS and SiB and is an absolute approach, where the magnitude of the stomatal resistance function is parameterized.

LEAF, on the other hand, adopted a relative approach, where the "potential evaporation" (maximum evaporation rate when there is no stomatal resistance) is evaluated first and then adjusted by a "relative stomatal conductance." This approach has been referred to as the "threshold concept," so that:

$$\lambda E_c = \rho \frac{c_p}{\gamma} \frac{d_{rs}}{\bar{r}_b} (e_a - e_*), \tag{10-21}$$

where d_{rs} is the relative stomatal conductance. This can be conceptually seen in Figure 10-1c, where the actual boundary aerodynamic resistance $\bar{r}_b$ is adjusted by a dial (i.e., d_{rs}).

Obviously, the stomatal resistance/conductance is still to be determined. Since transpiration is controlled by the stomata, a realistic parameterization of the stomatal opening is necessary in order to estimate the amount of latent heat flux correctly. Past studies showed that the stomata opening is affected by environmental variables (Allaway and Milthorpe, 1976; Jarvis, 1976; Avissar et al., 1985), and is parameterized in these models with the following forms:

BATS: $$r_s = r_{s\,\min} f_R f_S f_M f_V, \tag{10-22}$$

LEAF: $$d_{rs} = \frac{d_{\min} + (d_{\max} - d_{\min}) f_R f_T f_V f_\Psi}{d_{\max}}, \tag{10-23}$$

SiB: $$r_s = \left(\frac{a}{b + f_{\mathrm{PAR}}} + c\right) \frac{1}{f_L f_T f_V}, \tag{10-24}$$

where a, b, and c are plant-related constants, f_{PAR} is a environmental adjustment factor for photosynthetically active radiation (PAR), f_T is an adjustment factor for leaf temperature, f_V is an adjustment factor for vapor pressure deficit, f_R is an adjustment factor for total solar radiation, f_Ψ is an adjustment factor for soil water potential, f_S is an adjustment factor for seasonal temperature, and f_M is an adjustment factor for soil water availability. The subscripts min and max indicate the minimum and maximum values.

It is evident that both LEAF and SiB use functions of leaf temperature, soil water potential, solar radiation, and vapor pressure deficit. BATS considered radiation, water availability, seasonal temperature, and vapor pressure deficit. Note that BATS uses only one plant-dependent parameter, $r_{s\,\min}$, while LEAF and SiB use several. Although LEAF and SiB are more realistic in describing the stomatal response to environmental variables, a major difficulty in applying these models is to define these plant-related functions. Recently, Sellers et al. (1992) have modified the formulation of the stomatal conductance function to account for the rate of photosynthesis explicitly.

Model equations

If we can relate the resistance function to some environmental variables, as shown in the previous two sections, and assume the atmospheric condition is predicted, then the only unknown becomes the corresponding surface value. For example, the wind speed, air temperature, and water vapor pressure at the displacement height are required in order to estimate the amount of momentum, sensible heat, and latent heat fluxes. Since the surface roughness length and the displacement height are either calculated (SiB) or prescribed (LEAF and BATS), the momentum flux is immediately obtained by employing surface layer similarity theory. Sensible and latent heat fluxes, on the other hand, are still to be resolved. As mentioned before, the three models prescribe wind profiles in the canopy, which can either be a constant or varying with height. This will leave two variables to be determined, namely the water vapor pressure and the temperature of the canopy air (i.e., e_a and T_a in Figure 10-1). Assuming the canopy air has minimum heat or moisture storage, the turbulent heat fluxes gained from the soil and vegetation must be balanced by the loss to the atmosphere, so that:

$$H = H_c + H_{gs}, \text{ and} \tag{10-25}$$

$$\lambda E = \lambda E_c + \lambda E_{gs}. \tag{10-26}$$

Substituting Equations (10-11), (10-15), (10-16) and (10-20) or (10-21) into Equations (10-25) and (10-26), we can solve for the temperature (T_a) and the water vapor pressure (e_a) of the canopy air. However, we still need to determine the surface temperatures T_c and T_{gs}. This is done by solving the surface energy budget equation. Assuming a very small heat capacity in the canopy layer and the top soil layer, LEAF and SiB use a prognostic equation for the energy balance:

$$C_c \frac{\partial T_c}{\partial t} = Rn_c + H_c + \lambda E_c, \tag{10-27}$$

$$C_{gs} \frac{\partial T_{gs}}{\partial t} = Rn_{gs} + H_{gs} + \lambda E_{gs} + G, \tag{10-28}$$

where C_c and C_{gs} are heat capacity, in J m^{-2} K^{-1}, of the canopy and the top soil layer respectively. The variables Rn_c and Rn_{gs} are net radiation absorbed in the canopy and by the soil, and G is the ground heat flux to the deep soil layers. BATS, on the other hand, uses the balance equation at the vegetation surface:

$$0 = Rn_c + H_c + \lambda E_c\,, \tag{10-29}$$

and solves for the surface temperature, iteratively. For the soil surface, a "force–restore" method is used so that:

$$\frac{\partial T_{gs}}{\partial t} = \frac{c_1 G}{C_{gs}} - \frac{c_2(T_{gs} - T_{gz})}{\tau_1} \tag{10-30}$$

where c_1 and c_2 are constants, τ_1 is a time scale for soil heat transfer, and T_{gz} is a deep soil temperature. Solving the prognostic equations, as in LEAF and SiB, has the advantage of saving computation time but is less accurate and also introduces new parameters, namely the heat capacities C_c and C_{gs}. Fortunately these heat capacities are usually small so that the prognostic equations can still simulate the fast response of the surface temperature to the radiative forcing, especially if a very small time step (< 10 sec) is used as an example in LEAF. Using the iterative scheme not only increases the computational time requirement, but the model can fail to converge especially because the coupling between the surface and the atmosphere is a highly nonlinear process. It is also very well documented that soil heat and moisture transfers respond to temperature and soil moisture content in a very nonlinear manner (Clapp and Hornberger, 1978). The use of an iterative scheme with a soil model increases the chance of model failure.

Radiation fluxes

From Equations (10-27) and (10-28) or (10-29) and (10-30), it is obvious that the surface energy budget is mainly forced by radiation. A correct representation of the radiative flux in the canopy is necessary. The optical properties of the canopy are summarized by Sellers et al. (1986) so that three radiation bands are considered in SiB:

1. Visible or PAR (0.4 – 0.72 μm): most of the energy in this region is absorbed by the leaves.
2. Near infrared (0.72 – 4 μm): radiation is moderately reflective in this region.
3. Thermal infrared (> 4 μm): leaves behave like a black body in these wavelengths.

For the visible and near infrared regions, SiB treats direct and diffuse radiation separately. This is because the radiative transfer in the canopy for these short waves is highly dependent on the angle of the incident flux. For this reason, SiB also considers the change of surface albedo with solar angle, and the values are higher in the morning and evening when the solar angle is low. However, it might not be necessary to consider the variation of surface albedo since the optical depth also increases when albedo increases and the error in the net radiation should not be too large.

The surface albedo, in BATS and LEAF, is not varying with time, and there is no distinction between direct and diffuse radiation.

Soil representation

The major difference between the "greenhouse canopy" and the "big-leaf" model is that vegetation is treated separately from the ground surface in the first approach. In order to close the surface energy budget equations in the previous section [e.g., Equations (10-27), (10-28))], a soil model must be used to obtain the soil surface temperature T_{gs} and also the soil heat flux G. BATS and SiB use

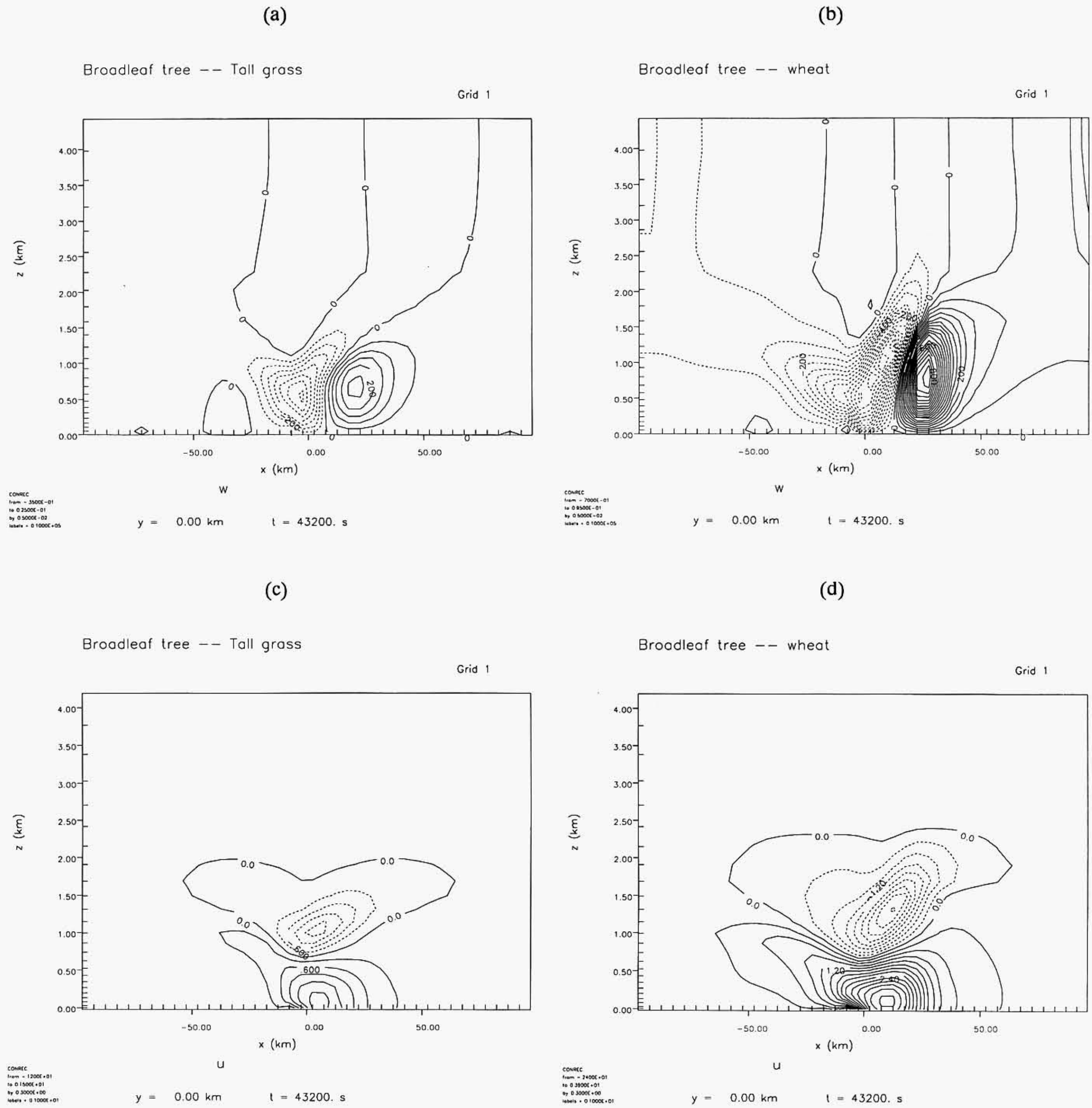
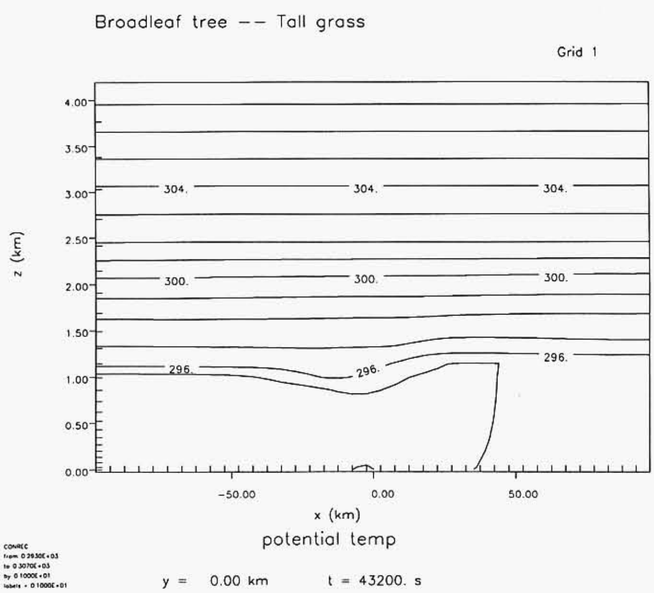
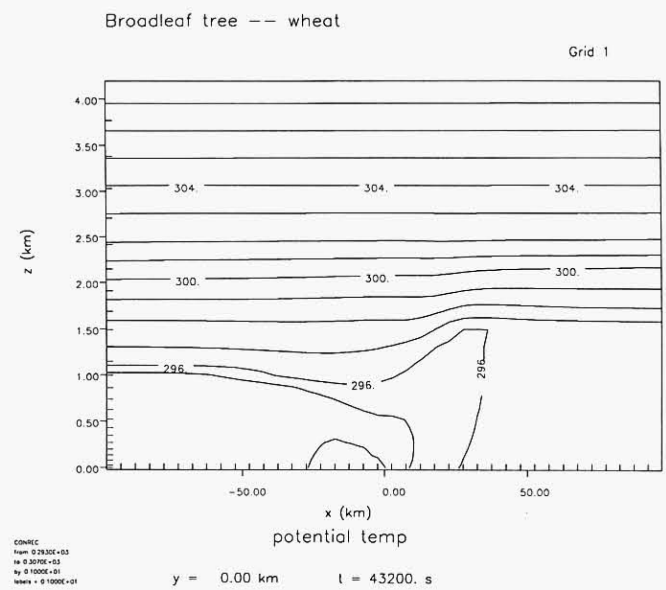


Figure 10-2: Atmospheric circulation at 1400 LST due to the contrast between a forest (left-hand side) and tallgrass prairie (left panels) and a forest and wheat farming (right panels). Contour intervals are (a) 0.5 cm s^{-1} for W; (b) 0.3 cm s^{-1} for U; (c)1 K for θ; and (d)1 g kg^{-1} for mixing ratio.

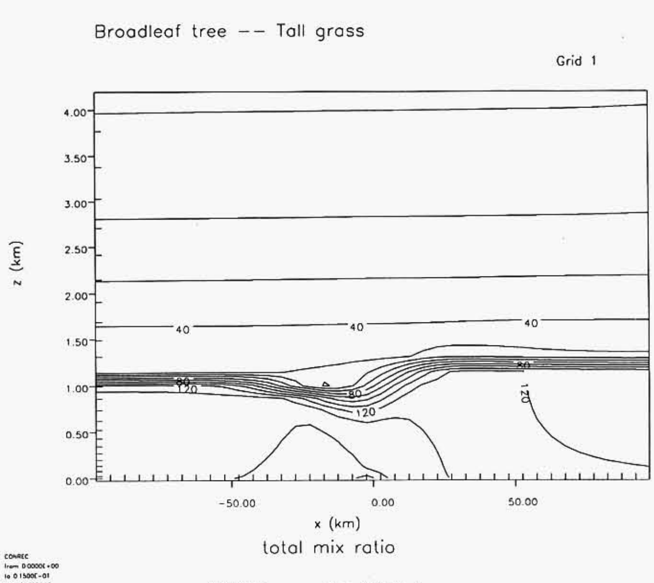
(e)


Broadleaf tree -- Tall grass
Grid 1
z (km)
x (km)
potential temp
y = 0.00 km
t = 43200. s

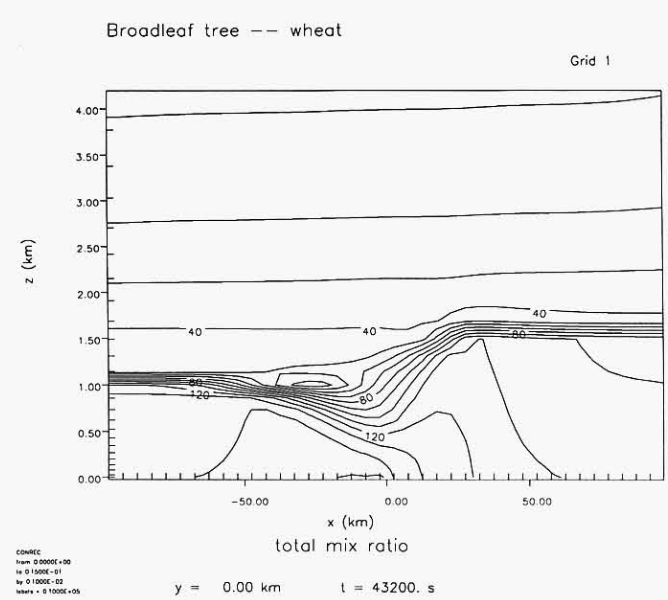
(f)


Broadleaf tree -- wheat
Grid 1
z (km)
x (km)
potential temp
y = 0.00 km
t = 43200. s

(g)


Broadleaf tree -- Tall grass
Grid 1
z (km)
x (km)
total mix ratio
y = 0.00 km
t = 43200. s

(h)


Broadleaf tree -- wheat
Grid 1
z (km)
x (km)
total mix ratio
y = 0.00 km
t = 43200. s

a force–restore method (Deardorff, 1978) that has three layers; LEAF uses a multilayer soil model (McCumber and Pielke, 1981) and has a detailed description for the transfer of moisture and temperature (Tremback and Kessler, 1985). The force–restore method is especially appropriate for climate models since it requires very little computer time. However, it fails to resolve the strong gradient of temperature and soil moisture potential close to the surface. Originally there were two soil layers used in the force–restore method such that the roots would be in the top soil layer with an averaged soil water content provided. However, the soil water potential can vary over several orders of magnitude from wet to dry so that it is very difficult to describe the appropriate average thermal and hydrological behaviors of the soil. It is also difficult to represent the soil water stress on vegetation correctly using this method. Due to the fact that this averaging process may lose important information on the water stress on vegetation, LEAF has chosen the use of a multilayer soil model. However, it is extremely difficult to initialize the multilayered soil model because of the lack of observations. Special measurements must be performed to obtain the needed information on soil moisture and temperature profiles.

Data requirement

In order to integrate an atmospheric model [e.g., Equation (10-1)] forward in time, initial and boundary conditions must be provided. For the initial conditions, an atmospheric model usually requires the state of atmosphere, which includes winds, temperature, pressure, humidity, and the state of soil, which includes soil moisture and temperature. For the surface boundary conditions, surface roughness and thermal and hydrological properties are essential. Soil texture, vegetation leaf area index, land-use, and land-cover data are necessary in order to estimate the surface thermal and hydrological properties. The percentages of surface coverage by different vegetation and landscapes are useful when a subgrid-scale land surface parameterization is used. DEM (digital elevation model) data can be used to define terrain characteristics. Clearly, it is the lower boundary condition information that GIS can contribute the most. In the future, it is expected that GIS can be linked to atmospheric modeling and analysis packages.

In this section, the major building blocks of three land surface parameterizations have been discussed. Many other model details are not covered in this chapter, for example, the treatment of dew, soil water flow, interception of precipitation by leaves, and evaporation from wet surfaces. Due to the fact that BATS and SiB are designed for use in GCMs, a complete hydrological cycle is available in these models. Both BATS and SiB can handle snow, ice, and dripping of water from leaves to the ground, while LEAF cannot. Generally SiB is more sophisticated and more realistic in the treatment of vegetation than BATS and LEAF. However, given the inhomogeneities in plant distribution and biophysical states within a model grid box (usually 400 × 400 km in GCMs and 20 × 20 km in RAMS), there is perhaps no need to use such a sophisticated model. Recently, Avissar (1992) proposed a statistical-dynamical approach in which the resistance functions are described by a distribution function, as an alternative to be used in numerical models.

THE IMPACT OF LANDSCAPE AND LAND USE ON ATMOSPHERIC CIRCULATIONS

Using these surface modeling schemes, the role of landscape type on planetary boundary layer structure and associated atmospheric circulations can be studied. Of specific interest is the difference in the partitioning of sensible and latent turbulent heat flux as a function of land use and landscape. Over irrigated areas and other areas of unstressed vegetation, boundary layer structure in the lower troposphere can be enhanced sufficiently to result in more vigorous cumulonimbus convection. Even slight differences in vegetation type, due to their different stomatal conductance and albedo characteristics, can cause substantial changes in the atmospheric response.

It has been shown in modeling (e.g., Ookouchi et al., 1984; Segal et al., 1988; Avissar and Pielke, 1989; Pielke and Avissar, 1990; Pielke and Segal, 1986) and observational studies (e.g., Segal et al., 1989; Pielke and Zeng, 1989; Pielke et al., 1990a) that the partitioning of sensible and latent heat fluxes into different Bowen ratios as a result of spatially varying landscape can significantly influence lower boundary layer structure and result in mesoscale circulations as strong as a sea breeze. Over and adjacent to irrigated land in the semiarid west, for example, enhanced cumulonimbus convection can result, as reported by Pielke and Zeng (1989). Schwartz and Karl (1990) document how the appearance of transpiring leaves on vegetation in the spring has the effect of substantially cooling (and thus moistening) the lower atmosphere. In their observational study, Rabin et al., (1990) demonstrate the effect of landscape variability on the formation of convective clouds. Dalu et al., (1991) evaluated, using a linear model, how large these heat patches must be before they generate these mesoscale circulations, while Pielke et al. (1991) present a procedure to represent this spatial landscape variability as a subgrid-scale parameterization in general circulation models.

These landscape variations result from a variety of reasons including:

1. Man-caused variations

- Agricultural practice (e.g., crop type, land left fal-

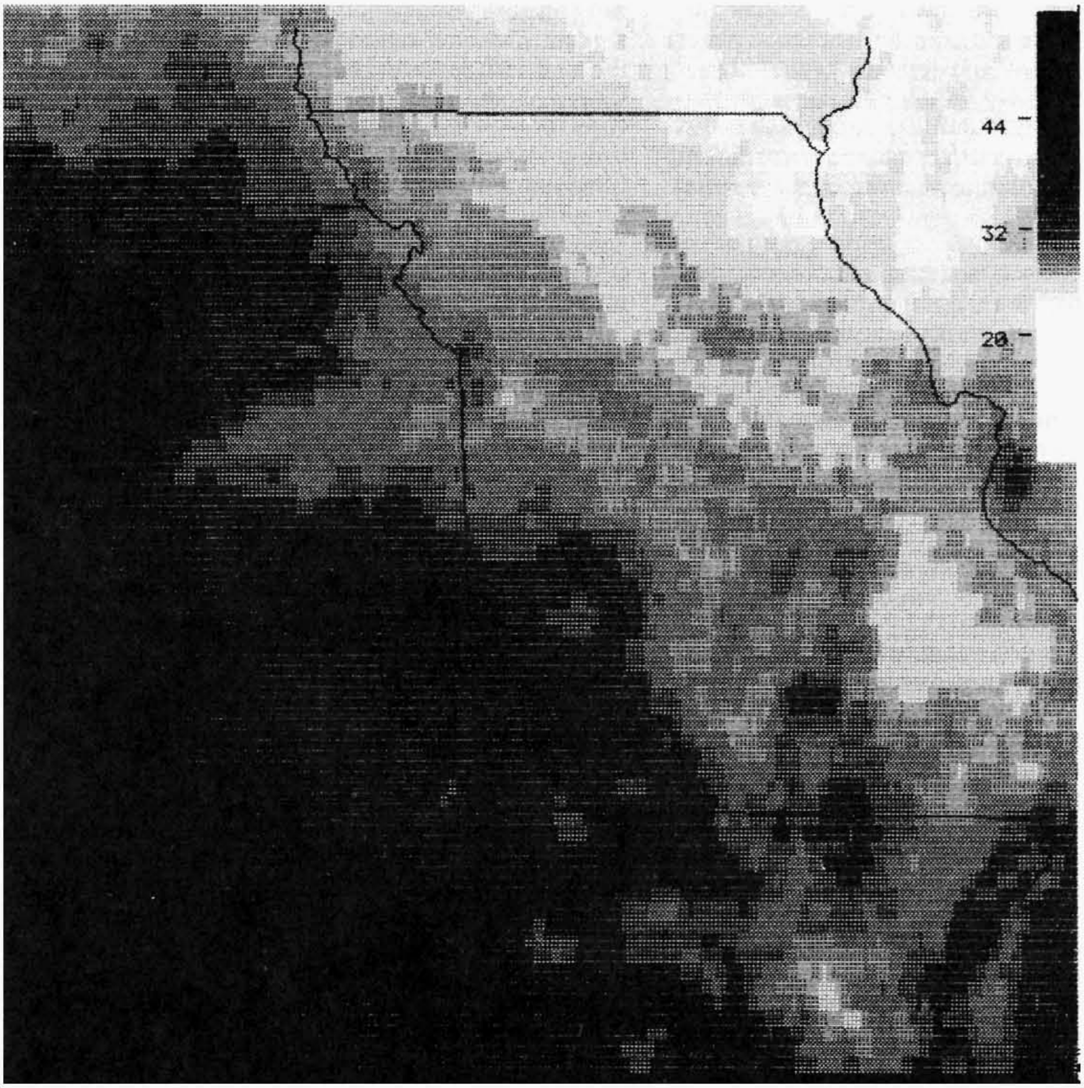


Figure 10-3: Plot of the highest measured surface skin temperature irradiance in °C as measured by the GOES geostationary satellite from July 24–August 31, 1986 for a region centered on eastern Kansas and western Missouri.

low, deforestation);

- Political practices of land subdivision (e.g., housing developments)
- Forest management (e.g., clearcutting);
- Irrigation.

2. Natural variations

- Fire damage to prairies and forests;
- Insect infestation and resultant damage to vegetation;
- Drought.

Plates 10-1 and 10-2 illustrate observed variations in photosynthetically active vegetation, as measured by NDVI (normalized difference vegetation index) satellite imagery over the Great Plains of the U.S. Presumably, active vegetation is transpiring efficiently during the daytime, while the other areas, with very low vegetation cover or vegetation under water stress, have most of their turbulent heat flux in sensible heat transfer. These two plates illustrate the large spatial and temporal variability of photosynthetically active vegetation and therefore suggest large corresponding variability in sensible heat flux (Pielke et al., 1991). GIS stored data must be of a sufficient spatial and temporal resolution to monitor these variations.

To illustrate the influence of landscape variations on weather, we have performed modeling and observational studies that demonstrate the major importance of vegetation and its spatial gradients on planetary boundary layer structure and mesoscale atmospheric circulations. Figure 10-2, for example, shows the results of a numerical model simulation with zero synoptic flow for the early afternoon in the summer for (1) a region in which a tallgrass prairie is adjacent to a forest region; and (2) the same as (1) except the tallgrass prairie is replaced by wheat.

For both simulations, the vertical velocity, east–west velocity, potential temperature, and mixing ratio fields are shown. Among the important results is the generation of a wind circulation as a result of the juxtaposition of the two vegetation types, and the change in the intensity of this circulation when the prairie is replaced by wheat. Higher transpiration over the forest, in conjunction with the thermally forced circulation, which can advect the elevated low-level moisture into the resultant low-level wind circulation, can be expected to result in enhanced convective rainfall when the synoptic environmental conditions are favorable. Changes in convective rainfall resulting from the conversion of the natural prairie to wheat also seem possible.

Satellite observations support the existence of large gradients in atmospheric conditions across a forest–grassland boundary in the United States, as illustrated in Figure 10-3 where the highest satellite-measured surface skin temperature irradiances are presented for a 5 week period in 1986 (July 24–August 31) as measured by the GOES geostationary satellite. Temperatures are over 10°C cooler over the forest as contrasted with prairie regions even short distances away (~30 km), as suggested to be the case by the modeling study.

CONCLUSIONS

Land-use and landscape patterns strongly influence atmospheric boundary layer structures and mesoscale circulations. For example, vegetation types as similar as tallgrass prairie and wheat cropland result in different atmospheric responses. This suggests that human modification of the land surface has had a major role in local climate, and, since such modifications have occurred worldwide, a global response to land-use changes should be expected. Through atmospheric modeling techniques, impacts of land-use change and its spatial heterogeneity on weather and climate systems can be studied and monitored. Consequently, there is a need for accurate characterization of the land surface for use as boundary conditions in atmospheric models. Important biophysical data for various vegetation species and hydrological data for soil states are necessary for correctly initializing these models. Clearly, given natural and human-induced spatial and temporal variability of land surface properties, model resolution needs to be greatly increased in order to include these features. GIS represents an important advancement in the refinement of global and regional land surface data sets that are necessary model inputs. Through improvements in model techniques and through the development of geo-referenced data sets, our future ability to model the global climate response to human-induced environmental change will be greatly enhanced.

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